

AD-A040 169

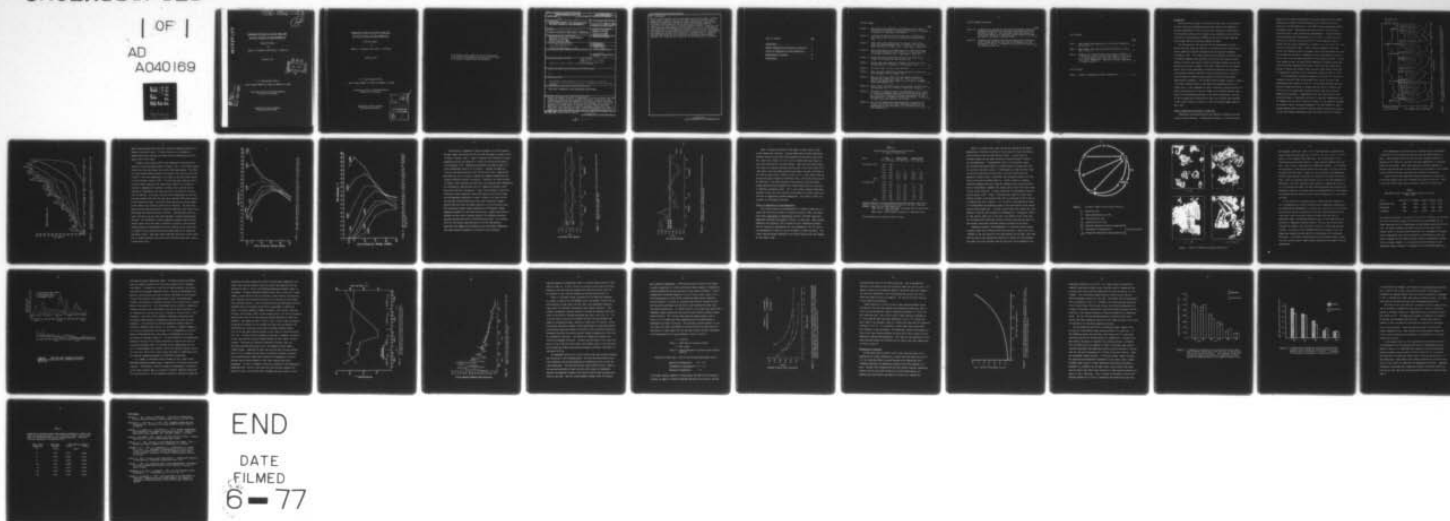
COLORADO UNIV BOULDER INST OF ARCTIC AND ALPINE RESEARCH F/G 8/12  
TEMPERATURE PATTERNS IN AN ALPINE SNOW COVER AND THEIR INFLUENC--ETC(U)  
FEB 77 E R LACHAPELLE, R L ARMSTRONG DAHC04-75-G-0028

UNCLASSIFIED

ARO-11866.1-6S

NL

| OF |  
AD  
A040169



AGRO-11866-1-G5 4

(12)

AD A 040 169

TEMPERATURE PATTERNS IN AN ALPINE SNOW COVER  
AND THEIR INFLUENCE ON SNOW METAMORPHISM

Technical Report

by

Edward R. LaChapelle and Richard L. Armstrong

February 1977



U. S. Army Research Office

Grants Number DAHC04 75 G 0028 and DAAG29 76 G 0088

Institute of Arctic and Alpine Research  
University of Colorado

Approved for public release;  
distribution unlimited.

AD NO. \_\_\_\_\_  
DDC FILE COPY

TEMPERATURE PATTERNS IN AN ALPINE SNOW COVER  
AND THEIR INFLUENCE ON SNOW METAMORPHISM

Technical Report

by

Edward R. LaChapelle and Richard L. Armstrong

February 1977

U. S. Army Research Office

Grants Number DAHC04 75 G 0028 and DAAG29 76 G 0088

Institute of Arctic and Alpine Research  
University of Colorado

Approved for public release;  
distribution unlimited.

ADDITIONAL for	
MTIS	White Section <input checked="" type="checkbox"/>
DUG	Buff Section <input type="checkbox"/>
UNANNOUNCED	<input type="checkbox"/>
JUSTIFICATION	
BY	
DISTRIBUTION/AVAILABILITY CODES	
Dist.	AVAIL. and/or SPECIAL
A	

The findings in this report are not to be construed  
as an official Department of the Army position, unless  
so designated by other authorized documents.

REPORT DOCUMENTATION PAGE		READ INSTRUCTIONS BEFORE COMPLETING FORM
1. REPORT NUMBER N/A	2. GOVT ACCESSION NO.	3. RECIPIENT'S CATALOG NUMBER
4. TITLE (and Subtitle) ⑥ TEMPERATURE PATTERNS IN AN ALPINE SNOW COVER AND THEIR INFLUENCE ON SNOW METAMORPHISM.		5. TYPE OF REPORT & PERIOD COVERED ⑦ Technical <sup>rept.</sup> Nov 1974-Nov 1976
7. AUTHOR(s) ⑩ Edward R./LaChapelle and Richard L./Armstrong		6. PERFORMING ORG. REPORT NUMBER
9. PERFORMING ORGANIZATION NAME AND ADDRESS Institute of Arctic and Alpine Research University of Colorado Boulder, Colorado 80302		8. CONTRACT OR GRANT NUMBER(s) ⑮ DAHC04-75-G-0028 DAAG29-76-G-0088
11. CONTROLLING OFFICE NAME AND ADDRESS U. S. Army Research Office Post Office Box 12211 Research Triangle Park, NC 27709		10. PROGRAM ELEMENT, PROJECT, TASK AREA & WORK UNIT NUMBERS
14. MONITORING AGENCY NAME & ADDRESS (if different from Controlling Office)		12. REPORT DATE ⑪ February 1977
		13. NUMBER OF PAGES ⑫ 42 p.
		15. SECURITY CLASS. (of this report) Unclassified
16. DISTRIBUTION STATEMENT (of this Report) Approved for public release; distribution unlimited.		15a. DECLASSIFICATION/DOWNGRADING SCHEDULE NA
17. DISTRIBUTION STATEMENT (of the abstract entered in Block 20, if different from Report) NA		
18. SUPPLEMENTARY NOTES The findings in this report are not to be construed as an official Department of the Army position, unless so designated by other authorized documents.		
19. KEY WORDS (Continue on reverse side if necessary and identify by block number) Snow cover; temperature; snow metamorphism; meteorology.		
20. ABSTRACT (Continue on reverse side if necessary and identify by block number) Spatial and temporal variations of temperature in alpine snow covers have been systematically observed over a period of two winters. Concurrently, snow crystal metamorphism has been monitored in the same snow covers, along with such basic snow properties as density and ram resistance. Near-surface snow temperatures fluctuate widely in response to diurnal weather variations. Below about 25 cm beneath the surface the temperatures change more slowly in response to longer-term weather trends. Mean snow temperatures are colder on north slopes than south ones but mean snow cover temperature gradients are		

DD FORM 1 JAN 73 1473

EDITION OF 1 NOV 65 IS OBSOLETE

Unclassified

SECURITY CLASSIFICATION OF THIS PAGE (When Data Entered)

406 875 CT

## 20. (continued)

similar on both exposures owing to shallower snow on south slopes. A forest canopy tends to suppress snow surface radiation cooling and hence reduce magnitude of temperature gradients at depth. Metamorphism in snow follows a recrystallization mode with declining mechanical strength when the saturation water vapor pressure gradient exceeds 0.05 mb/cm. Owing to a non-linear vapor pressure-temperature relationship over ice, this corresponds to the conventional critical temperature gradient of  $0.1^{\circ}\text{C/cm}$  for this metamorphism mode only at snow temperatures close to the melting point. Mean monthly snow temperature gradients can reasonably be estimated from air temperature and snow depth means, but this method can be extended to vapor pressure gradients only if appropriate corrections for non-linearity are introduced.

Unclassified

TABLE OF CONTENTS	<u>Page</u>
<u>Introduction.....</u>	1
<u>General Temperature Distribution in Snow Cover....</u>	1
<u>Effects of Temperature on Snow Metamorphism.....</u>	11
<u>Meteorological Influences.....</u>	26
<u>Bibliography.....</u>	33

## LIST OF FIGURES

	<u>Page</u>
Figure 1. Near-surface snow temperatures at Red Mountain Pass (3400 m), showing typical patterns of temperature fluctuations influenced by external air temperature and snowfalls.....	3
Figure 2. Isotherm distribution within the snow cover at Red Mountain Pass (3400 m) for the period 15 October 1974 through 30 April 1975.....	4
Figure 3. Hourly near-surface temperatures of snow on a north-facing slope near the Red Mountain Pass site (3400 m) for the period 0830 to 1430 MST on 30 March 1975. Slope angles is about 30°..	6
Figure 4. Hourly near-surface snow temperatures on a south-facing slope near the Red Mountain Pass site (3400 m) for the period 0830 to 1430 MST on 30 March 1975. Slope angle is about 25°.....	7
Figure 5. Average snow cover temperatures on north- and south-facing slopes for mid-January to late February 1975.....	9
Figure 6. Average snow cover temperature gradients on north- and south-facing slopes for mid-January to late February 1975.....	10
Figure 7. The major stages of snow crystal evolution.....	14
Figure 8. Mean snow cover temperature gradients measured at three sites from 1 November 1974 to 30 March 1975.....	18
Figure 9. Mean ram resistance, density and mean monthly temperature gradients for a depth hoar layer (H = 20-40 cm) in the snow cover at Red Mountain Pass (3400 m) for the period 1 November 1974 to 28 February 1975.....	21
Figure 10. Winter trend of saturation water vapor pressure gradient across the same snow layer (H = 20-40 cm ) identified in Figure 9.....	22
Figure 11. The degree of recrystallization (TG metamorphism) within layers of new snow accumulating under natural conditions on bare ground as a function of saturation vapor pressure gradient and time. Each data point is the mean of several observations. A total of 48 observations were made over two winters.....	25
Figure 12. The critical temperature gradient required to establish the critical saturation water vapor temperature gradient for TG metamorphism in snow, 0.05 mb/cm, as a function of mean snow layer temperature.....	27

## LIST OF FIGURES (continued)

	<u>Page</u>
Figure 13. A comparison of observed and calculated mean monthly temperature gradients through the entire snow cover at Red Mountain Pass for the winter of 1974-75. The calculated values were obtained by dividing mean monthly air temperature by mean monthly snow depth according to the method of Akitaya (1974).....	29
Figure 14. A comparison of observed and calculated vapor pressure gradients (mean) through the entire snow cover at Red Mountain Pass for the winter of 1974-75. The method for calculating the gradients is explained in the text.....	30

## LIST OF TABLES

	<u>Page</u>
Table 1. Near-surface Snow Temperatures as a Function of Amount of Forest Cover.....	12
Table 2. Mean Monthly Vapor Pressure Gradients (mb/cm) at 3 Sites 1974 - 1975.....	17
Table 3. Comparison of saturation water vapor pressure gradients in a 100-cm snow cover for the two snow layers $H = 0-10$ cm and $H = 0-100$ cm (entire snow cover) for different values of snow surface temperature. Snow-earth interface temperature is assumed to be $0^{\circ}\text{C}$ .....	32

## LIST OF PLATES

Plate I. Stages of temperature-gradient metamorphism.....	15
-----------------------------------------------------------	----

## Introduction

During the winter seasons of 1974-1975 and 1975-1976, an investigation into the radiation and temperature conditions governing the temperature-gradient metamorphism of snow, with particular reference to deterioration of load-bearing capacity of the snowcover, was undertaken by the Institute of Arctic and Alpine Research (INSTAAR) University of Colorado under ARO Grants Number DAHC 04-75-G-0028 and DAAG29 76 G 0088.

The area chosen for the study was the San Juan Mountains of southwestern Colorado, where the character of the radiation snow climate with extensive temperature-gradient metamorphism had already been documented (LaChapelle, 1974). The primary objective of the research was twofold; one, to produce fundamental data regarding the nature of the recrystallization process in natural snow covers by spatial and temporal mapping of snow temperature and structural regimes in the local climate, and two, to develop a body of data on exact magnitudes of temperature gradients and associated vapor pressure gradients required to induce recrystallization, as well as practical means of predicting these gradients from readily observable meteorological parameters. The following Technical Report contains a summary of the most important results of the above study presented primarily in graphic form. A more comprehensive report containing a detailed analysis of data acquired during this study will appear as an INSTAAR Occasional Paper. Details of the methodology employed in the field investigations, the type of data collected and a description of field site locations were presented in the original proposal as well as in the first Progress Report dated 30 April 1975.

## General Temperature Distribution in Snow Cover

Temperatures were measured within the snowcover throughout the winter season at three elevations. Thermocouples attached to a fixed array were

spaced at 10 cm intervals beginning at the ground surface at the Silverton-Lakawanna site (2850 m) and the Chattanooga site (3160 m) and at 20 cm intervals at the Red Mountain Pass site (3400 m) where considerably greater snow depths prevail. Measurements were made just prior to sunrise on a daily basis with a Leeds and Northrup potentiometer. The calibration of the potentiometer was checked weekly in an ice bath. In addition to the fixed array, a portable array consisting of 5 rigid probes was used to measure the near-surface temperatures at depths of 2.5, 5.0, 10.0, 15.0, 20.0 and 25.0 cm below the surface. Figure 1 gives an example of such near-surface snow temperature data for November 1974 at the Red Mountain Pass site. Also included are minimum air temperature values on the morning of observation as well as the new snow accumulation of the preceding 24 hours. During clear weather periods, the near-surface temperatures respond to prevailing surface radiation temperatures consistently lower than those found during periods of overcast and snowfall, when the near-surface layers reflect the air temperature conditions of the storm period, generally  $0.0^{\circ}\text{C}$  to  $-5.0^{\circ}\text{C}$ .

A time-stratigraphic diagram of temperature variations ( $^{\circ}\text{C}$ ) within the snowcover during 1974-1975 at Red Mountain Pass, one of the three fixed thermocouple sites, appears in Figure 2. The isotherms represent that temperature regime which exists far enough below the snow-air interface (25-35 cm) so as to be appreciably insulated from the short-term of diurnal temperature fluctuations. Temperatures within this lower portion of the snowcover respond to longer-term variations in mean daily temperature, with the response-time lag being a function of depth. As the snowcover continues to increase in depth, the general tendency is for the isotherms to slowly migrate upwards seeking to maintain the same distance from the snow surface. A significant warming trend began during early March and the  $0^{\circ}\text{C}$  isotherm

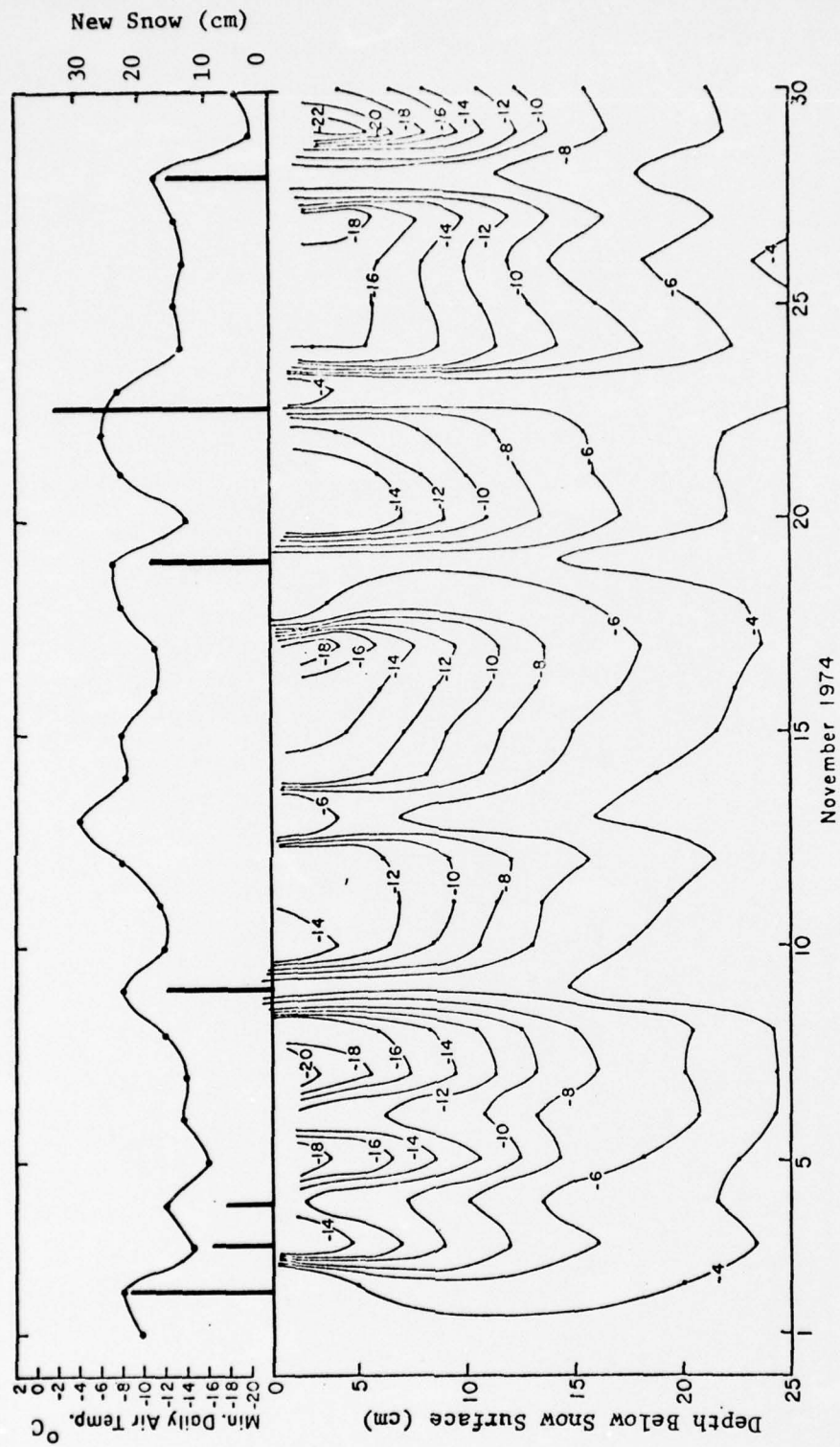


Figure 1 Near-surface snow temperatures at Red Mountain Pass (3400 m), showing typical patterns of temperature fluctuations influenced by external air temperature and snowfalls.

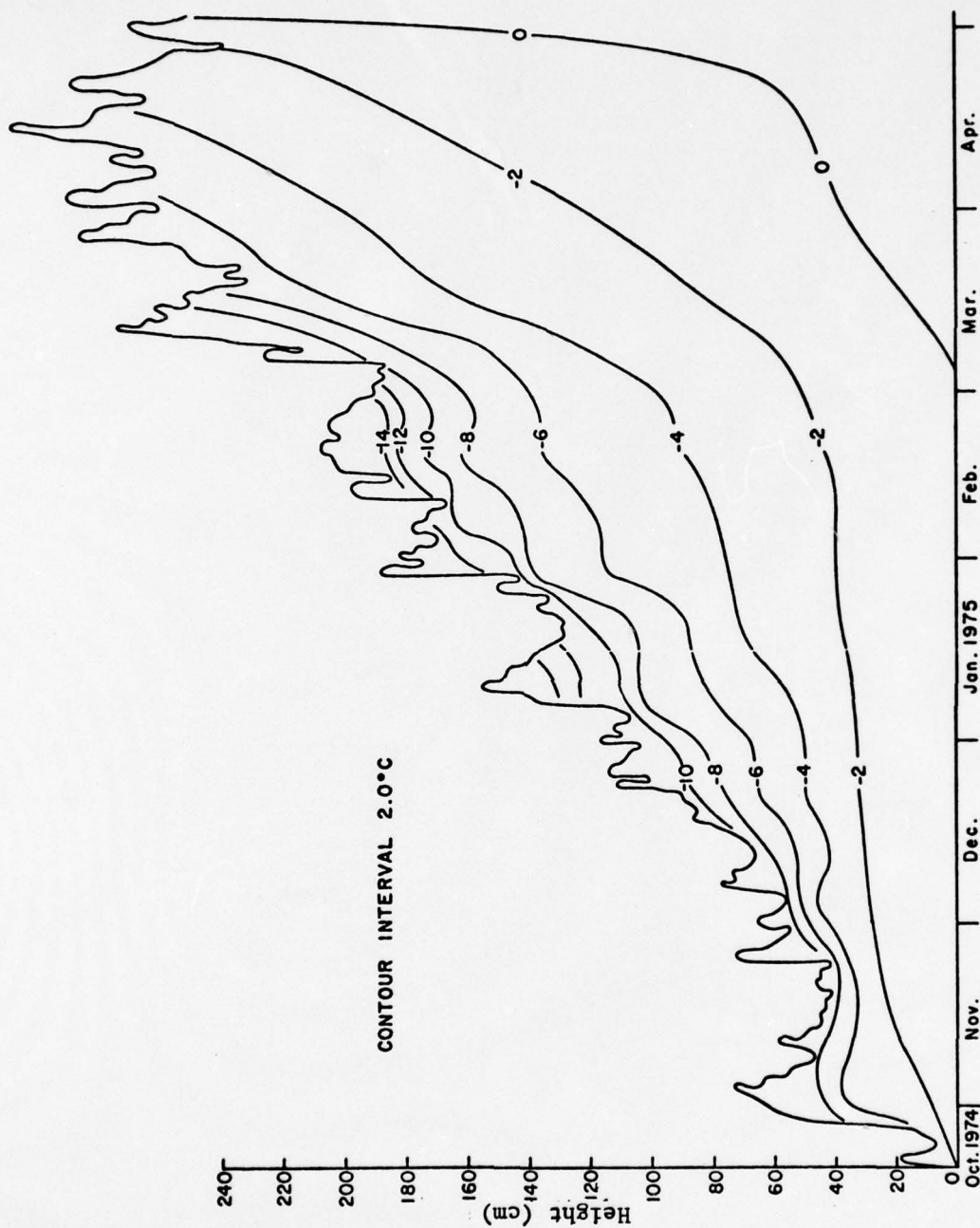


Figure 2 Isotherm distribution within the snow cover at Red Mountain Pass (3400 m) for the period 15 October 1974 through 30 April 1975.

began to move towards the surface until the entire snowcover becoming isothermal by the end of April. A further discussion of isotherm and temperature-gradient patterns associated with each thermocouple site will appear later in this report.

Examples of near-surface diurnal snow temperature fluctuations with respect to slope orientation appear in Figures 3 and 4. The portable thermocouple array, described above, was used for these measurements. The effect of diurnal warming nearly vanishes at the depth of 25 cm on the north slope (Figure 3), while the temperature at this depth on the south slope (Figure 4) varied through a range of  $3.0^{\circ}\text{C}$ . It can be seen that during early morning on the south slope while the layers above a depth of 10 cm were increasing in temperature in response to incoming solar radiation and increasing air temperatures, the layers below 10 cm continued to lose heat until mid-morning. By mid-day, the near-surface temperatures on the south slope had reached  $0.0^{\circ}\text{C}$  while the same layers remained  $5.0^{\circ}\text{C}$  below freezing on the north-facing slope. The north-facing slope was *not* totally shaded at this time but was receiving indirect solar radiation. The south slope began receiving unobstructed direct radiation immediately after sunrise with the optimum solar angle occurring at 1300 MST. Snow densities within the layers involved at the two sites were comparable, averaging approximately  $150 \text{ kg/m}^3$ . The difference in diurnal temperature regimes within the near-surface layers of the two slopes is primarily a function of the efficient warming produced by the penetration of solar radiation on the south slope, in contrast to heat received by the north slope primarily via conduction from the air above. These data clearly show that the effect of solar radiation is largely confined to the top 25 cm of the alpine snow cover, even on a south-facing slope.

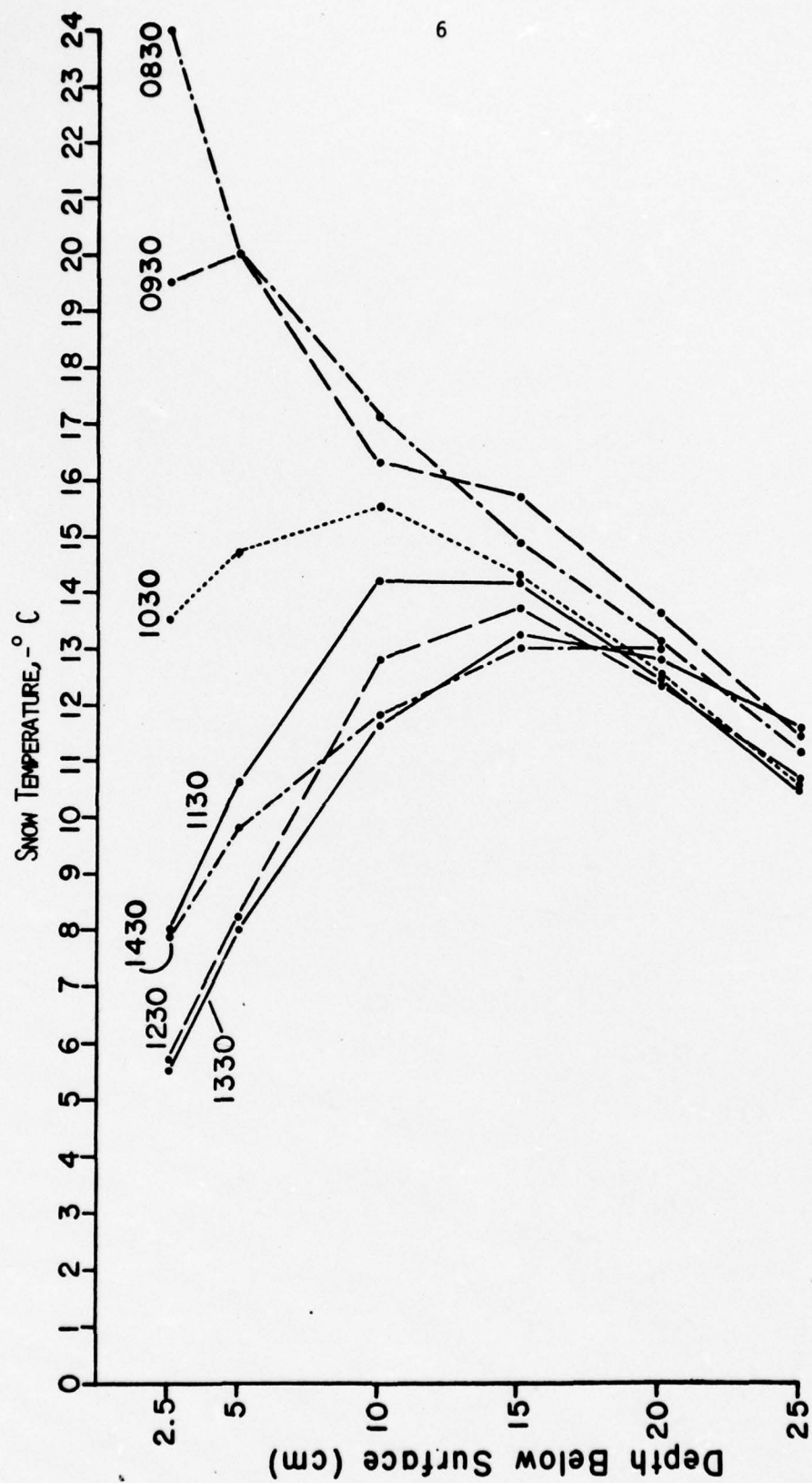


Figure 3 Hourly near-surface temperatures of snow on a north-facing slope near the Red Mountain Pass site (3400 m) for the period 0830 to 1430 MST on 30 March 1975. Slope angles is about  $30^{\circ}$ .

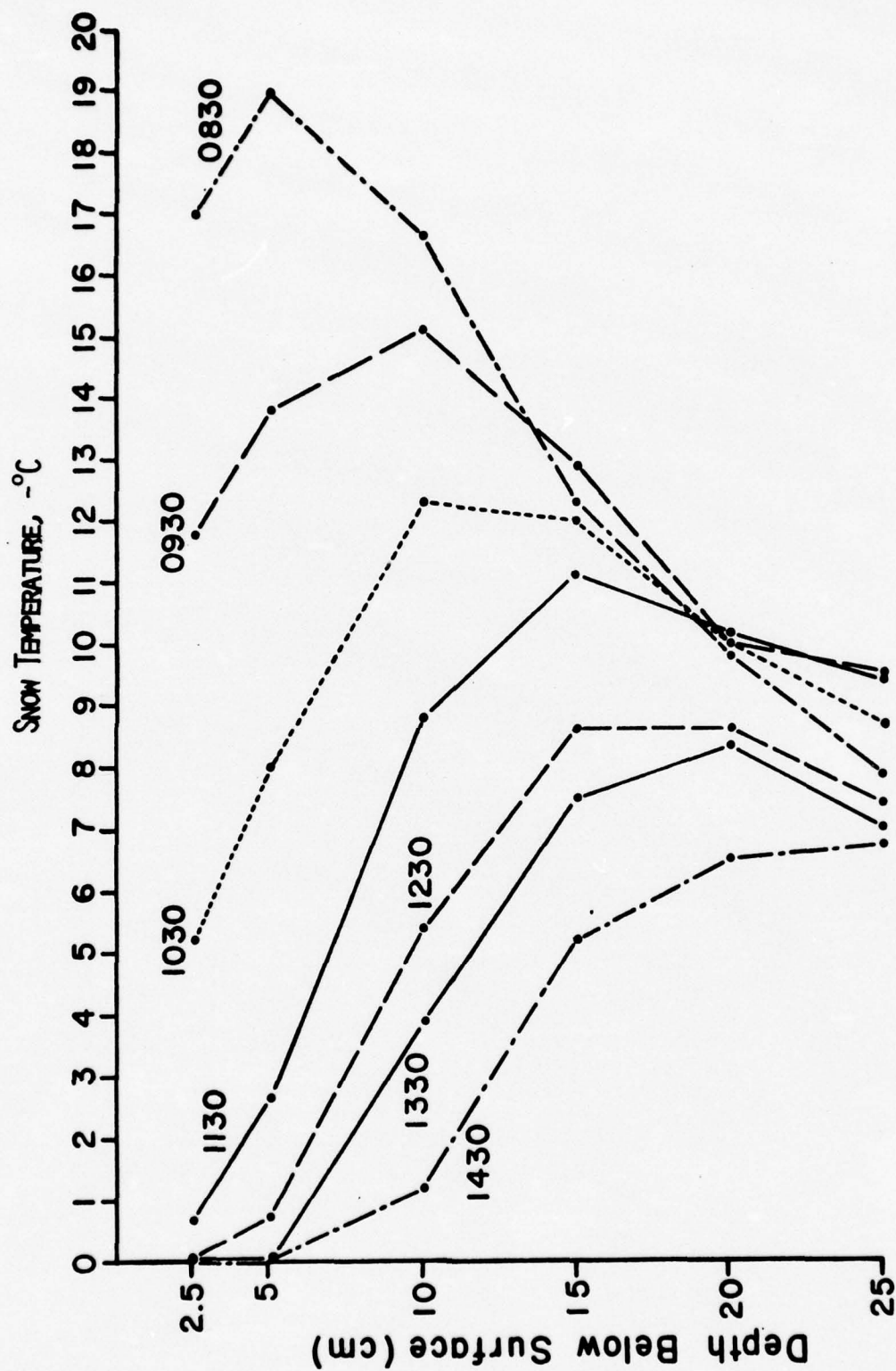


Figure 4 Hourly near-surface snow temperatures on a south-facing slope near the Red Mountain Pass site (3400 m) for the period 0830 to 1430 MST on 30 March 1975. Slope angle is about 25°.

The variation in temperature regimes throughout the entire snowcover and over longer time periods for the two sites described in Figures 3 and 4 is shown in Figures 5 and 6. Figure 5 indicates the difference in average temperatures within the snowcover as a whole for the period mid-January to late-February 1975. Temperature measurements were made by means of a single thermocouple probe at 20 cm intervals. Average snow depth was 1.50 m for the south slope and 1.80 m for the north slope. Observations were made just prior to sunrise, dropping the uppermost temperature most strongly affected by diurnal fluctuations. Therefore, the values plotted in Figures 5 and 6 represent that portion of the snowcover where temperatures are influenced by long-term mean daily air temperature patterns rather than short-term daily fluctuations. Figure 5 does indicate a significant difference in mean temperature values from south to north-facing slopes but the temperature gradients, as shown in Figure 6, are quite similar, with the more shallow south-facing slope often indicating a steeper gradient than the colder but deeper snowcover on the north-facing slope. With consistently higher average temperatures and frequently steeper temperature gradients on the south-facing slope, a higher vapor pressure gradient can thus create a greater potential for temperature-gradient recrystallization than exists on the north-facing slope. This finding, valid for the high-altitude conditions of the San Juan Mountains, contradicts the commonly-held assumption that the strongest temperature (and vapor pressure) gradients are confined to north exposures.

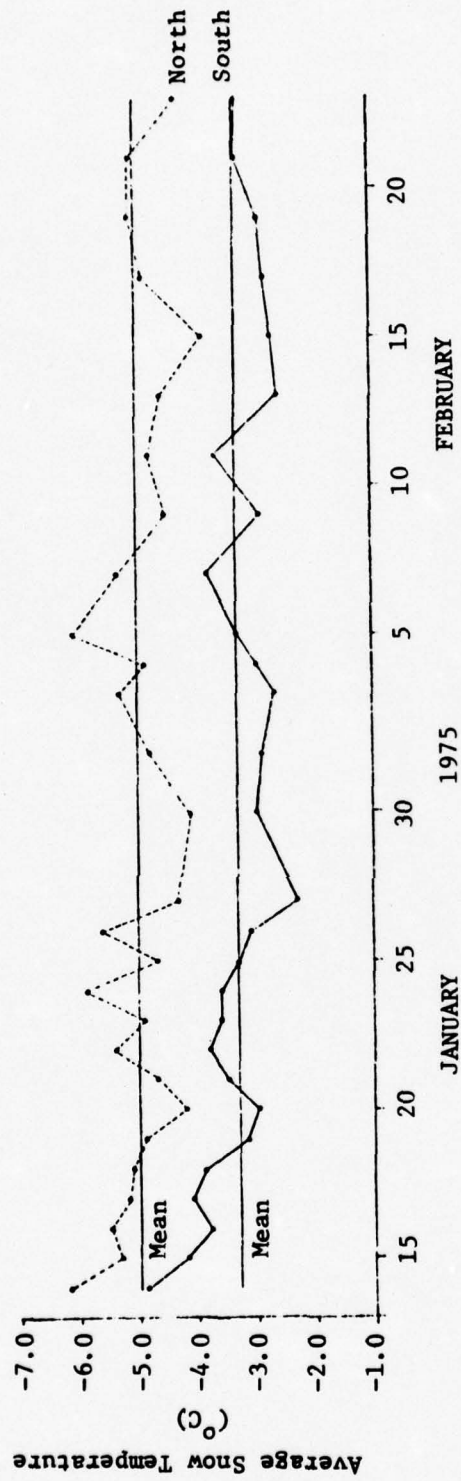


Figure 5 Average snow cover temperatures on north- and south-facing slopes for mid-January to late February 1975.

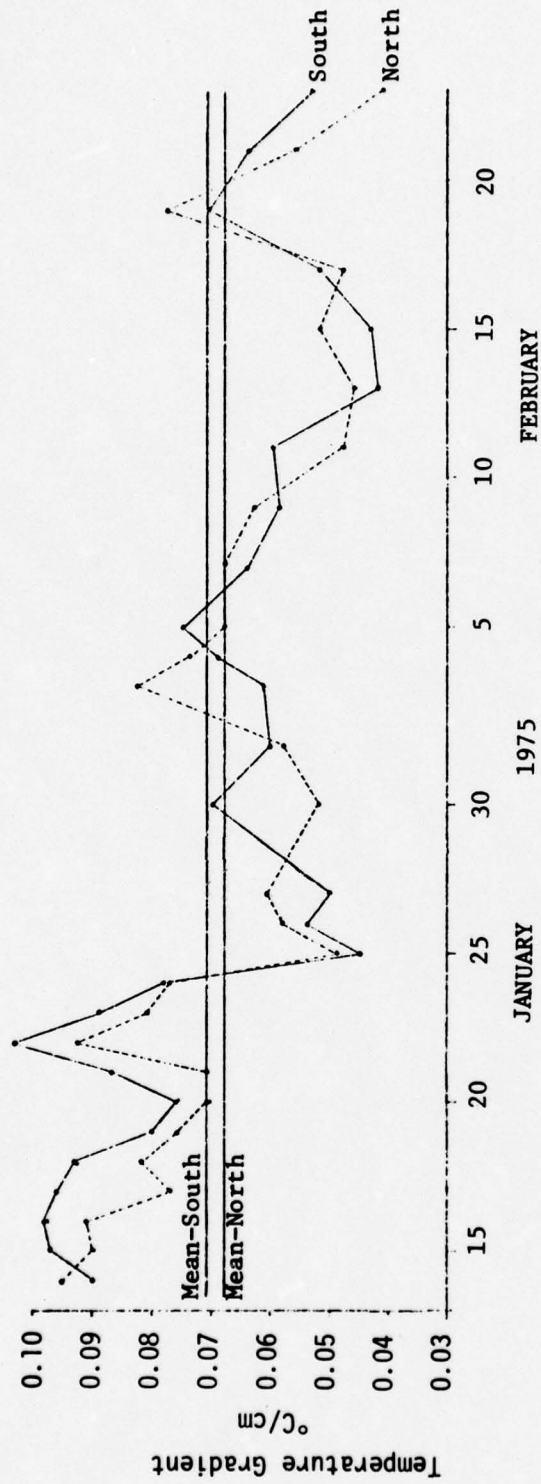


Figure 6 Average snow cover temperature gradients on north- and south-facing slopes for mid-January to late February 1975.

Table 1 indicates the effect of the amount of forest cover on near-surface temperature conditions. Average temperatures and mean temperature gradients were calculated from data provided by the portable surface array for a layer from a depth of 2.5 cm to 25.0 cm beneath the snow surface at sites with three distinct radiation regimes. Open sites were unobstructed with respect to outgoing long-wave radiation with no tree cover within 20 m. Light forest sites were unobstructed directly above, but were surrounded on all sides by trees within a radius of from 4 to 6 m. Heavy forest sites had the sky obscured by branches of coniferous trees. The salient conclusion to be drawn here is that open exposure of snow surfaces to the sky permits strong radiation cooling compared with locations where the sky is partially or completely obscured by trees. This in turn induces stronger temperature gradients at the clear site, pointing to the effect of a forest canopy as an inhibitor of temperature-gradient metamorphism. This effect is found consistently in both February and March.

#### Effects of Temperature on Snow Metamorphism

Two principal modes of snow metamorphism are commonly recognized, following the distinction originally introduced by Eugster (1952), who termed these modes constructive and destructive according to whether appreciable temperature gradients were present or absent in a snow layer. More recently, Sommerfeld and LaChapelle (1970) proposed the term "temperature-gradient" (TG) for constructive metamorphism and "equi-temperature" (ET) for destructive metamorphism in order to clarify the genesis of these two modes. This latter usage has become widespread in the United States and has been adopted for the present study.

Table 1

Near-surface Snow Temperatures as a Function  
of Amount of Forest Cover

Date	Open		Light Forest		Heavy Forest	
	$\bar{T}$ °C	$\overline{TG}$ °C/cm	$\bar{T}$ °C	$\overline{TG}$ °C/cm	$\bar{T}$ °C	$\overline{TG}$ °C/cm
26 February 1975	-16.0	.458	-	-	-	-
	-15.8	.396	-	-	-	-
	-14.9	.409	-	-	-	-
	-15.1	.413	-12.2	.276	-10.8	.249
	-16.0	.360	-12.2	.240	-10.6	.222
	Mean	-15.6 .407	-12.2 .258	-10.7 .236		
19 March 1975	-10.6	.462	-	-	-	-
	-10.2	.458	-9.3	.345	-7.4	.151
	-10.4	.427	-9.5	.373	-7.6	.196
	-10.8	.427	-8.5	.271	-7.5	.187
	-10.4	.431	-8.5	.289	-7.7	.156
	-9.9	.400	-8.6	.311	-6.9	.191
	Mean	-10.4 .434	-8.9 .318	-7.4 .176		

Average temperature and mean temperature gradient through a layer from a depth of 2.5 cm to 25 cm beneath the surface at sites with 3 radiation regimes: Open = no tree cover within 20 m

Light Forest = site clear above, surrounded 360° by trees within radius of 4-6 m

Heavy Forest = site covered by branches of coniferous trees

All measurements were made just before sunrise.

Layers in a natural winter snow cover may be subjected to one mode of metamorphism or the other according to the dictates of climate and progressive evolution of the snow cover during a winter. A given layer may alternate between the two modes according to external weather influences on snow temperatures. ET metamorphism leads to a fine-grained, dense, well-sintered snow whose mechanical strength normally increases with time and can reach substantial values. TG metamorphism at lower densities leads to recrystallized snow with large grains, inferior sintering and low mechanical strength, hence it diminishes the load-bearing capacity of snow, often to the point of extreme fragility. Figure 7 presents a schematic convention adopted during this study for representing the various possible paths of metamorphism, together with symbols used to depict the major stages of snow crystal evolution. A similar convention has independently been proposed by Pahaut (1975). Plate I illustrates four stages of crystal metamorphism relevant to the present study, which concentrated on the TG mode as it affected snow cover structure. I-A, I-B and I-C show progressive stages of TG metamorphism from beginning stages of recrystallization to the advance stage of mature depth hoar. I-D shows crystals which have reached the intermediate TG stage and then reverted to ET metamorphism. Photographic records of snow crystals types such as this were a key feature of this study which was introduced after preliminary efforts to obtain consistent and objective descriptions from several different observers proved to be fruitless.

Temperature-gradient (TG) metamorphism is a recrystallization process induced by water vapor diffusion within the snow cover. Because the solid framework of the snow consists of the same substance as the vapor, the latter does not need to flow through the connected air spaces over long distances, but rather can simply sublime from one snow grain and re-condense on the

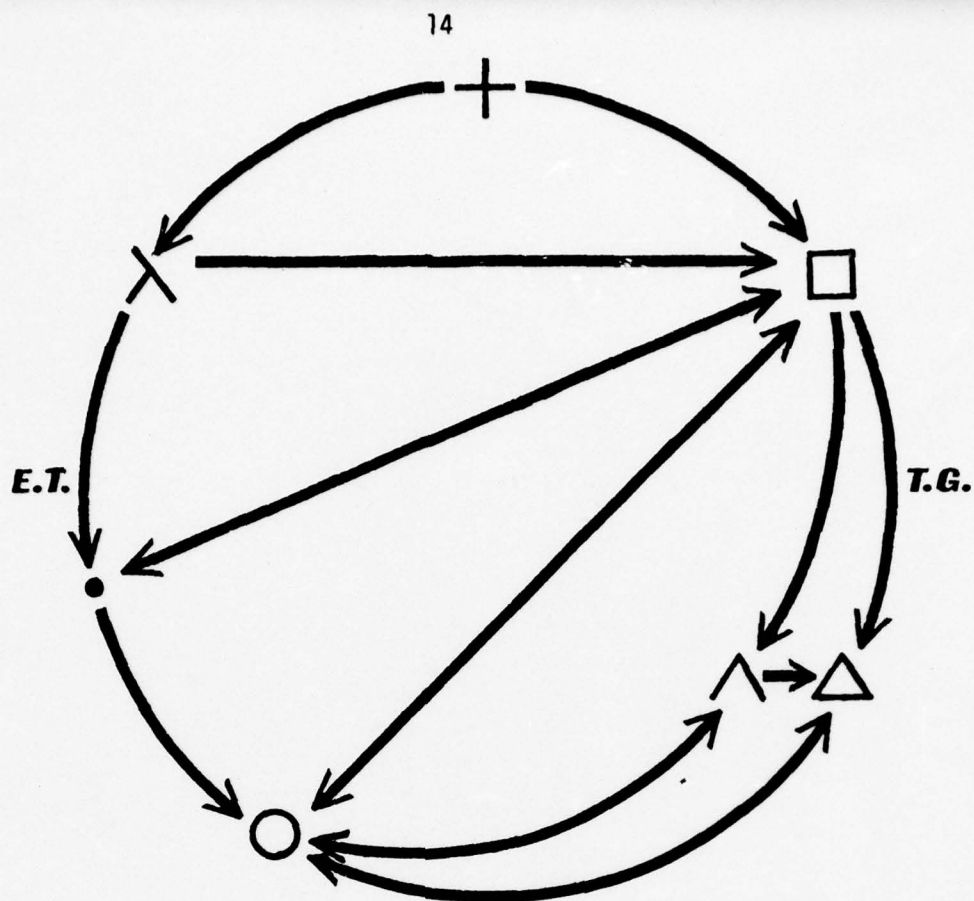


Figure 7      The major stages of snow crystal evolution.

- + -new-fallen snow
  - ∧ -partly-metamorphosed new snow
  - -fine-grained old snow
  - -coarse-grained old snow
  - -beginning temperature-gradient metamorphism
  - ∧ -intermediate TG metamorphism
  - △ -advanced TG metamorphism (mature depth hoar)
- ] recrystallization



I-A □



I-B ^



I-C △



I-D ^ → ○

Plate I Stages of temperature-gradient metamorphism

next adjacent, colder one. Much or all of the snow layer in question thus passes through the vapor phase and is re-formed into new crystals, called TG snow or, in its completed form, depth hoar. The driving force for this diffusion/recrystallization process is a vapor pressure gradient within the snow. This in turn is established by a temperature gradient. If the reasonable assumption is made that the intercrystalline spaces in a snow layer are saturated with water vapor in respect to ice at the prevailing temperature, then the vapor pressure at any point can readily be calculated from the well-known values of saturation vapor pressure over ice, once the temperature is determined. Since this saturation vapor pressure-vs.-temperature curve for ice is distinctly non-linear, the vapor pressure gradient which controls the metamorphic processes in the snow cover can be determined only if the temperature gradient and the absolute value of the mean temperature in a snow layer are known.

In the course of his pioneering work on snow structures, Eugster (op. cit.) recognized that a critical minimum value of temperature gradient must be exceeded before metamorphism in snow goes over from the ET to the TG mode, but he did not specify just what this value was, perhaps because he recognized that it was not an exact figure but rather could vary over an appreciable range. Accumulated experience since his time consistently has pointed to a temperature gradient around  $0.1^{\circ}\text{C}/\text{cm}$  as the critical value, although this number is not sufficient to specify a unique vapor pressure gradient. As recently as the comprehensive work by Akitaya (1974), the temperature gradient has been taken as the determining factor for snow metamorphism mode. It has been a primary objective of the present study to establish the critical value of vapor pressure gradient which divides ET from TG metamorphism.

Snow temperatures are relatively easy to measure and have constituted a large part of the field observations upon which the present analysis is based. Vapor pressures within the snow cover are extremely difficult to measure but can readily be calculated from known temperatures if saturation conditions are assumed. Water vapor pressures and derived gradients of pressure reported here have all been determined by calculation in this fashion and are referred to below as observed vapor pressures and gradients.

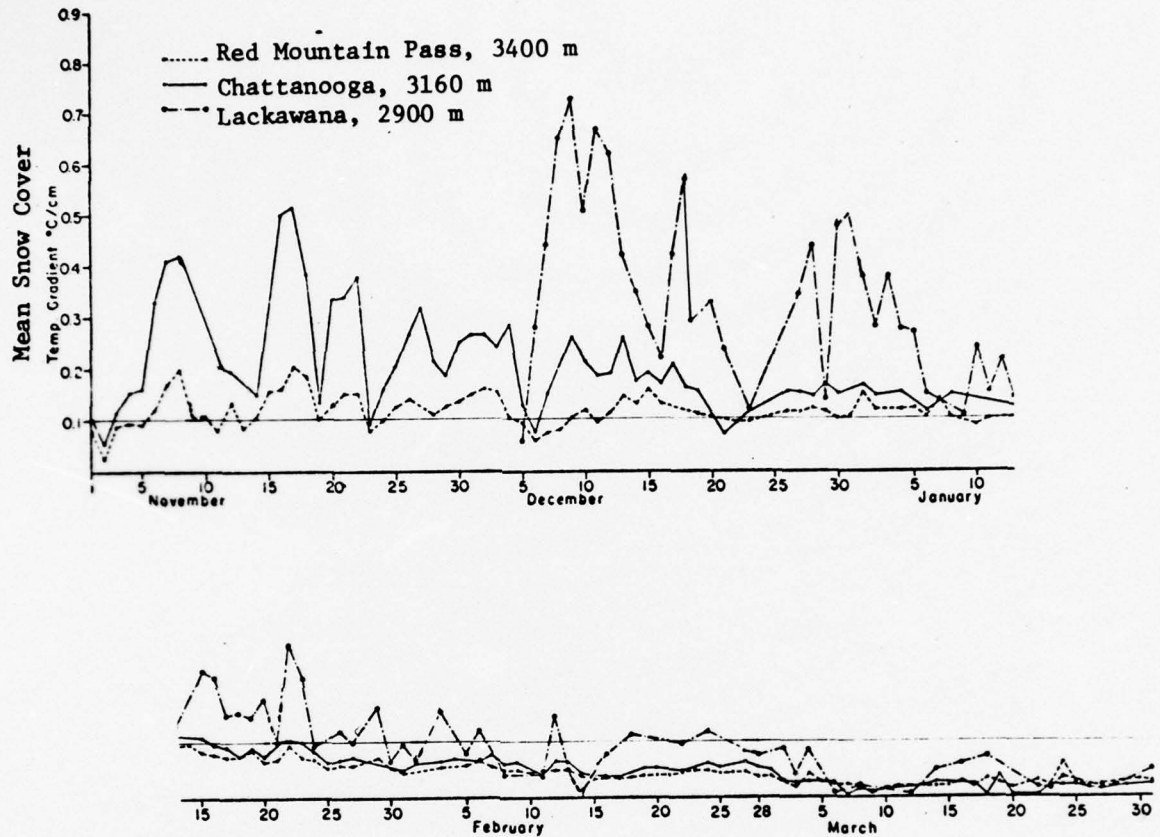
Figure 8 exhibits the temperature gradients measured on a daily basis at the three primary study sites throughout the winter of 1974-1975. Table 2 gives mean monthly vapor pressure gradient values for the same sites.

Table 2

Mean Monthly Vapor Pressure Gradients (mb/cm) at 3 Sites  
1974 - 1975

Site	Elevation (m)	December	January	February	March
Red Mountain Pass	3400	.0454	.0315	.0137	.0118
Chattanooga	3160	.0704	.0407	.0239	.0099
Lackawanna	2850	.1501	.0679	.0354	.0162

The temperature gradients, as measured at three sites throughout the snowcover just prior to sunrise by means of the fixed array, are dependent on the longer range mean daily air temperatures and snow depths associated with each site. The general trends at the three sites are similar, with initial gradient values in the early winter being high, as a consequence of frequent low air temperatures and a shallow snowcover, and diminishing as the snow height increases and late winter and early spring air temperatures increase. The early season, however, is also a period of wide fluctuation in the temperature values, especially in November prior to the establishment of a



**Figure 8** Mean snow cover temperature gradients measured at three sites from 1 November 1974 to 30 March 1975.

consistent mid-winter temperature regime. The lowest average snow temperatures and steepest gradients were consistently measured at the Lakawanna site (2850 m). Although this is the site of lowest elevation, two factors contribute to the extreme temperature values. Because the measurements are made at a valley station just prior to sunrise, the effect of the nocturnal valley inversion during clear weather greatly lowers snow temperatures. Secondly, the snowcover is relatively shallow at this location with an average depth of less than 0.8 m. Also, the greatest daily variation in temperature gradient occurred at this low elevation site, due to wide fluctuations in air temperature and shallow snowcover as compared to the two sites at higher elevations. The valley site at Lakawanna consistently experienced lower minimum and higher maximum air temperatures than the two sites at higher elevations. At the three sites the mean temperature gradients began to diminish at different times from mid- to late-winter. Gradients dropped to  $0.1^{\circ}\text{C}/\text{cm}$ , the conventional cut-off point for TG metamorphism, during mid-January at Red Mountain Pass, late January at Chattanooga and not until early March at Lakawanna (Figure 8). The early reduction of the temperature gradient at the two sites at higher elevations is primarily the result of increasing snow depth, even at a time when the air temperatures are still low. At the Lakawanna site, however, because the snowcover remains shallow, it is not until the arrival of early spring and higher air temperatures that the snowcover temperature gradient is significantly diminished.

It is common in continental alpine climates to find a layer of well developed temperature-gradient snow, or depth hoar, at the base of the winter snowcover. The thickness, relative strength and topographical distribution of this snow structure type is a function of climatic conditions during the fall and early winter. As the snowcover continues to increase in height and

subsequent snow layers become more distant from the warmer temperatures and higher vapor pressure gradients nearer the ground, the opportunity for the formation of thick ( $> 10$  cm), well-developed temperature-gradient layers is diminished. The basal temperature-gradient layer, which in the local climate commonly varies from 20 to 60 cm in thickness, often provides a failure zone which undermines the load-bearing capacity of the snow and also facilitates later avalanche release. A few investigations have been undertaken previously regarding physical conditions controlling the development of depth hoar layers in a natural snowcover (Trabant and Benson, 1972; Bradley, Brown and Williams, 1976; Akitaya, 1974), but the present study is the first to establish quantitative values of the vapor pressure gradient necessary for depth hoar formation. One example of this is found in Figs. 9 and 10, where the density and mechanical strength of a 20-cm depth hoar layer near the bottom of the Red Mountain Pass snow cover are plotted for November through February (Fig. 9) and the vapor pressure gradient of this layer for the entire winter (Fig. 10). Following initial settlement of new snow in November, the density of this layer declines slightly through December and then remains virtually constant. The mechanical strength as measured by the mean layer ram resistance declined precipitously until late December and then began a steady increase. Comparison of these data with the vapor pressure gradient trend in Fig. 10 shows that the decline in mechanical strength associated with recrystallization (depth hoar formation) was predominant as long as the mean vapor pressure gradient of this layer remained above about 0.05 mb/cm. The rise in strength, reflecting a reversion to equi-temperature metamorphism and sintering, took place only after the mean gradient fell below this value. By the third week in December the snow crystals in this

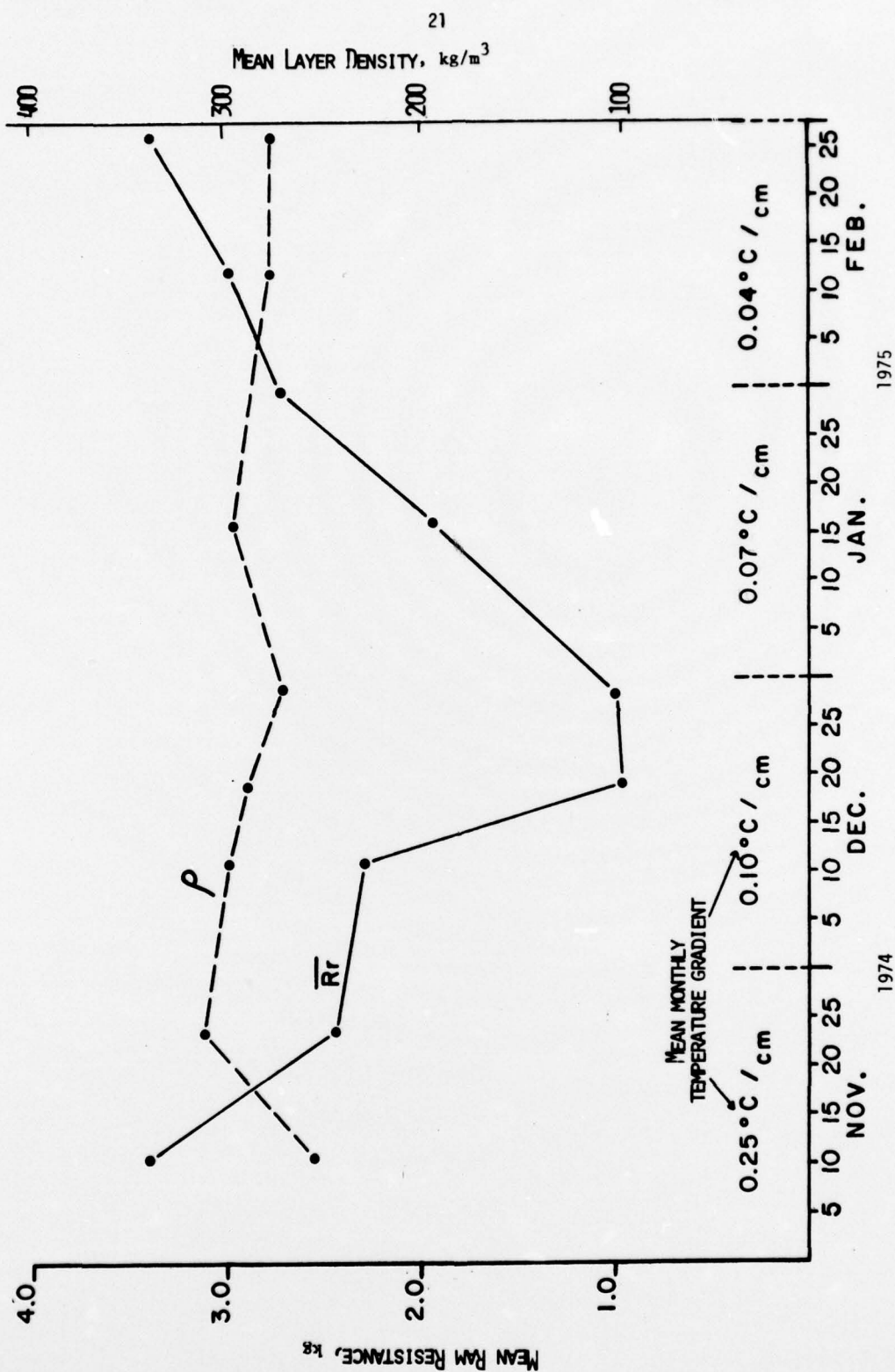


Figure 9 Mean ram resistance, density and mean monthly temperature gradients for a depth hoar layer ( $H = 20\text{-}40\text{ cm}$ ) in the snow cover at Red Mountain Pass (3400 m) for the period 1 November 1974 to 28 February 1975.

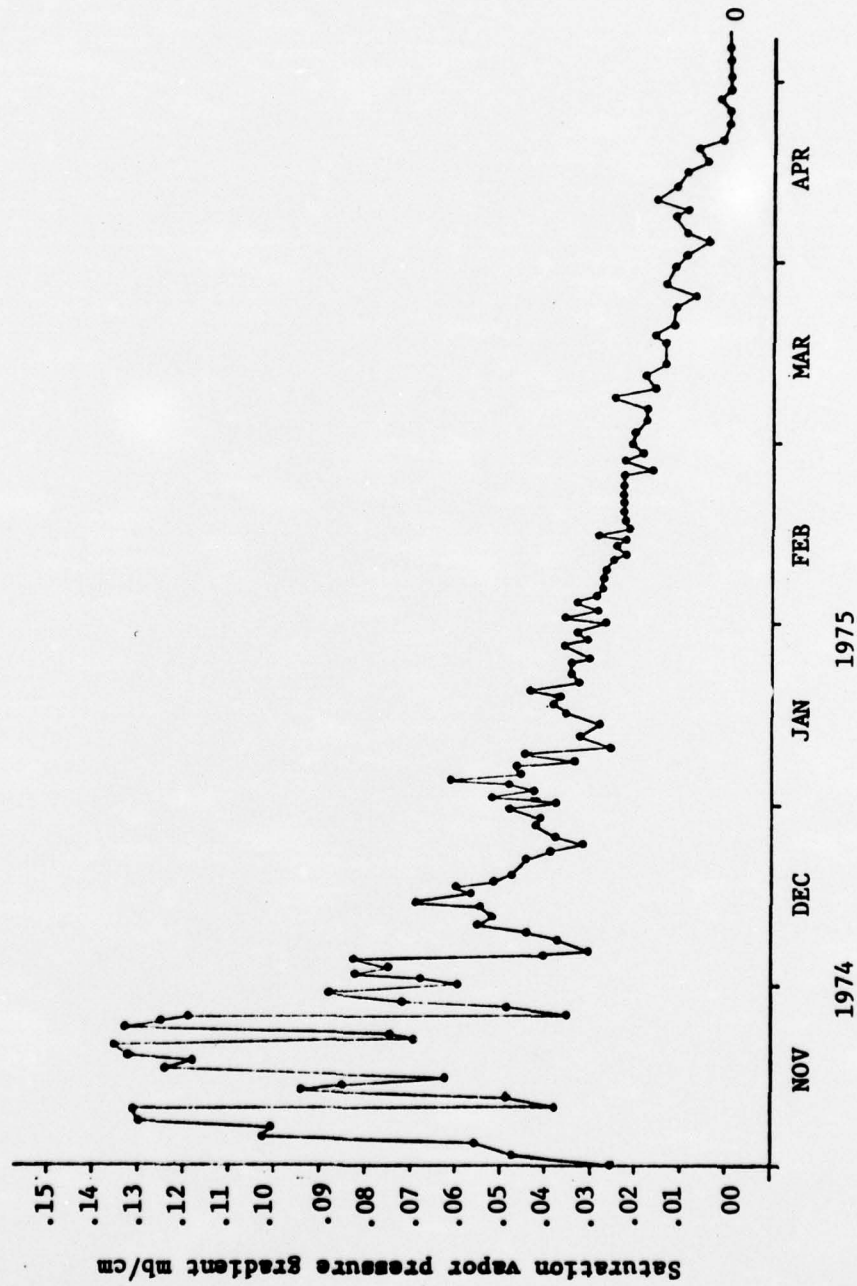


Figure 10 Winter trend of saturation water vapor pressure gradient across the same snow layer ( $H = 20-40$  cm) identified in Figure 9.

layer had reached an intermediate stage of recrystallization typical of that shown in Plate I-B. By early January the recrystallization process had been reversed as the vapor pressure gradient fell below the critical level and the crystals in this layer had assumed the form typified by Plate I-D.

Figure 9 illustrates further characteristics of depth hoar formation. The decline in density from mid-November until late December reflects the upward migration of water vapor from this layer into subsequently deposited layers under the influence of persistent vapor pressure gradients. Under extreme circumstances such mass depletion can erode the lowermost snow cover layers to the point of collapse (see Bucher and others, 1940, Fig. 10). The recrystallization process also increases the grain size and decreases the number of intergranular bonds, leading to a sharp rise in viscosity. Unlike fine-grained snow which responds to each loading event (accumulated snowfall) by densifying, depth hoar remains sufficiently stiff to minimize compressive deformation and hence the density persists largely unchanged even at the base of a substantial snow cover. The mechanical strength also remains relatively low throughout the winter. The mean ram resistance of this layer rose only to 13 kg at the end of the winter, while higher layers in the snow cover which escaped the intense vapor pressure gradients achieved a final mean ram resistance of 54 kg.

An independent test for the critical value of the vapor pressure gradient was carried out in the following manner. At sites receiving minimal direct solar radiation, the existing snowcover was removed just prior to each new snow accumulation. The resulting shallow, usually 20-50 cm thick, layers of new snow were monitored at least once daily with respect to temperature gradient and metamorphic changes, while density profiles were recorded every three to four days. Specific crystallographic changes within the samples

were recorded by photography. Sufficient data were acquired by this method to permit recognition of a clear relationship between progress of temperature-gradient metamorphism and the magnitude of vapor pressure gradients. Figure 11 parameterizes the observed degree of recrystallization within samples of new snow (average density 50-150 kg/m<sup>3</sup>) accumulating under natural conditions and bare ground as a function of saturation vapor pressure gradient and time. Each data point associated with the three curves represents the mean value from several observations. In Figures 9 and 10, it was noted that the decrease in ramsonde strength ceased once the prevailing vapor pressure gradient dropped below 0.05 mb/cm. This critical vapor pressure gradient appears again in Figure 11 as a minimum value which apparently must be exceeded before TG metamorphism can begin in new snow. Once the vapor pressure gradient in snow rises above this value, the degree of recrystallization can approximately be described by a simple product of time and vapor pressure gradient represented by the experimental set of quasi-hyperbolic curves in Figure 11.

$$k = t (\text{grad } P)$$

where  $k$  = coefficient of recrystallization

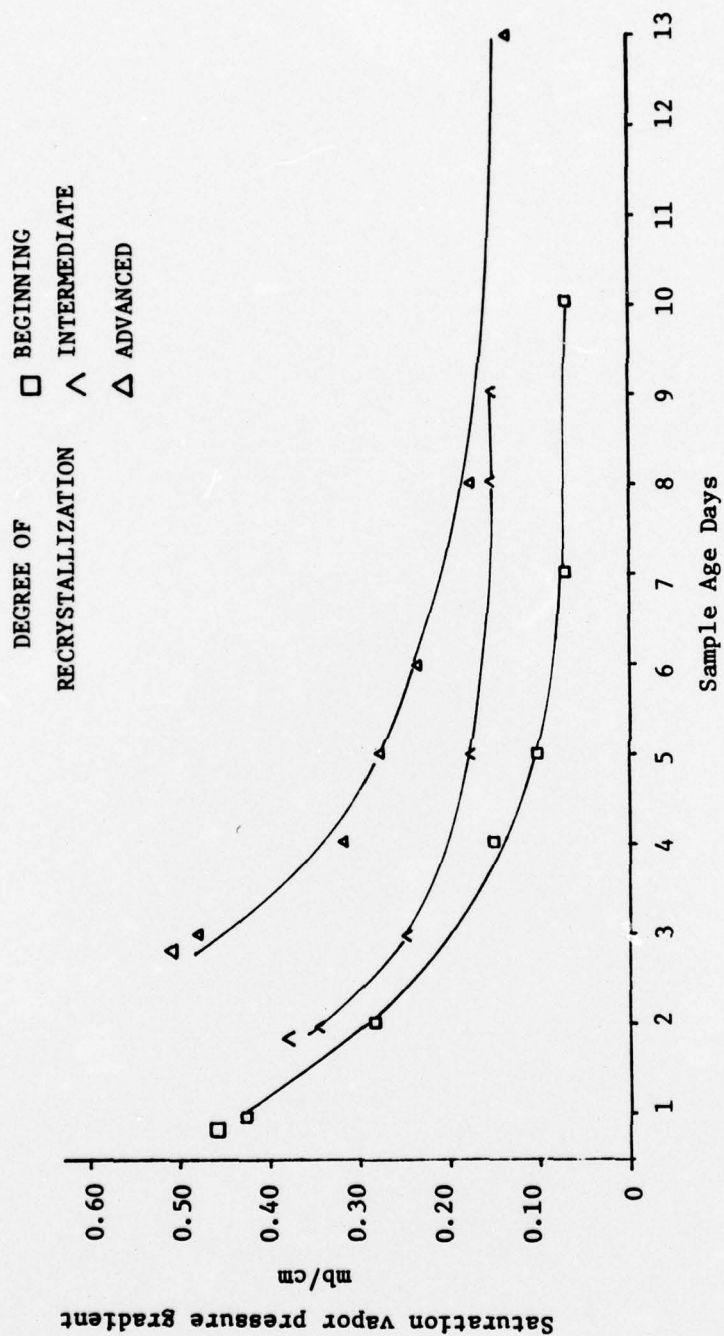
$t$  = time in days

(grad  $P$ ) = scalar magnitude of saturation vapor pressure gradient

According to these data,  $k$  then has the following approximate values:

beginning TG metamorphism	0.4 - 0.6
intermediate TG metamorphism	0.7 - 1.3
advanced TG metamorphism	1.4

If the vapor pressure gradient is large enough, the effects of TG recrystallization can appear in naturally deposited new snow in as quickly as one day



**Figure 11** The degree of recrystallization (TG metamorphism) within layers of new snow accumulating under natural conditions on bare ground as a function of saturation vapor pressure gradient and time. Each data point is the mean of several observations. A total of 48 observations were made over two winters.

and mature depth hoar can form within three days. One to two weeks are required if the gradient rises only marginally above the critical value. The range of  $k$ -values for intermediate degree of recrystallization are probably the least well determined in this study because the crystals fall into a wider type range difficult to categorize. The initial and final forms can be more readily distinguished.

Identification of a critical value of vapor pressure gradient around 0.05 mb/cm as the cross-over between snow metamorphism modes gives some insight into the conventional figure of temperature gradient,  $0.1^{\circ}\text{C}/\text{cm}$ , for this same cross-over. The critical value of vapor pressure corresponds to  $0.1^{\circ}\text{C}/\text{cm}$  when the mean snow layer temperature is 0 to  $-1^{\circ}\text{C}$ , values commonly found in the lowermost layers of a winter snow cover when the snow-earth interface is at  $0^{\circ}\text{C}$ . It is precisely in these lower layers where depth hoar formation is most pronounced. The temperature gradient corresponding to the critical vapor pressure gradient of course has in fact a range of values depending on ambient snow temperature. This range for the determined critical vapor pressure gradient of 0.05 mb/cm and for typical snow cover temperatures is given in Figure 12.

#### Meteorological Influences

As previously seen in Figures 3 and 4, short term variations in air temperature are largely reflected in a surface snow layer about 25 cm thick. It is the long-term effect of prevailing mean daily temperatures which ultimately determines the temperature gradient within the snowcover as a whole. Akitaya (1974) suggested that the mean monthly snowcover temperature gradient could be calculated indirectly by dividing mean monthly air temperature by mean monthly snow depth at a given site, assuming the

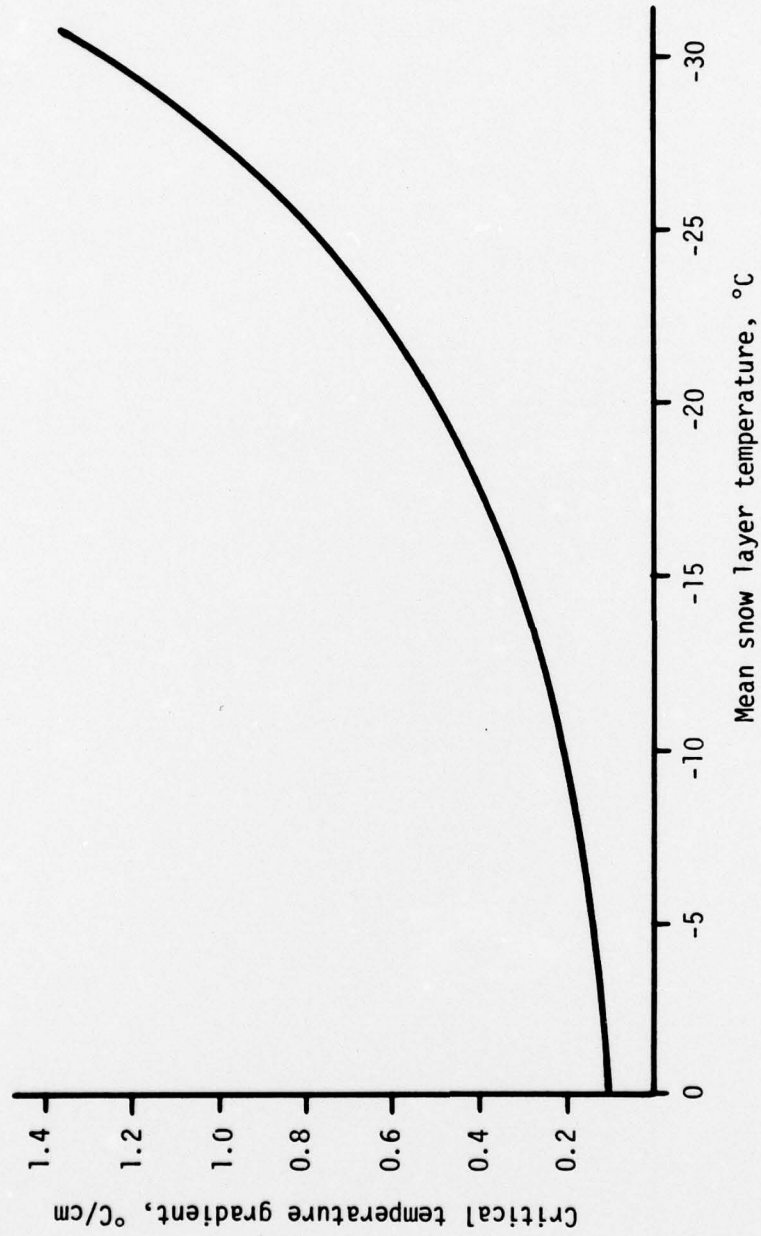


Figure 12 The critical temperature gradient required to establish the critical saturation water vapor temperature gradient for TG metamorphism in snow, 0.05 mb/cm, as a function of mean snow layer temperature.

snow-ground interface to be at  $0^{\circ}\text{C}$ . This study, having collected daily snowcover temperature-gradient values, snow depth and air temperature data at three sites, had an excellent opportunity to check the accuracy of such a method. The relationship between Akitaya's indirect calculation and direct measurement proved to be very good. One example from the Red Mountain Pass site appears in Figure 13. Although it has been shown that especially in this high altitude, low latitude climate, snow surface temperatures are strongly influenced by incoming short wave and out-going long wave radiation conditions, such radiation balances do indirectly establish air temperature values. The fact that the extreme temperature variations within near-surface snow layers apparently balance out over longer periods of time allows the application of the method suggested by Akitaya.

We have tested the possibility of extending Akitaya's method to the calculation of mean monthly vapor pressure gradients in a snow cover. If the difference between the saturation vapor pressure at  $0^{\circ}\text{C}$  (snow earth interface) and that for the mean monthly air temperature is divided by the mean monthly snow depth, an apparent vapor pressure gradient is obtained. Such calculated values are compared in Figure 14 with mean vapor pressure gradients determined from temperature profiles actually observed in the snow cover for the same circumstances as in Figure 13 (see also Table 2). Again, the two methods compare favorably. A difficulty arises, however, because the vapor/pressure-temperature relationship is non-linear, hence a simple arithmetic mean based on a linear temperature distribution with height, a reasonable approximation for the lower layers of most alpine snow covers, does not reflect the rather large variations of vapor pressure gradient with height in such a snow cover. Thus, in Figure 13 the nominal critical temperature gradient of  $0.1^{\circ}\text{C}/\text{cm}$  is exceeded by the monthly snow cover mean

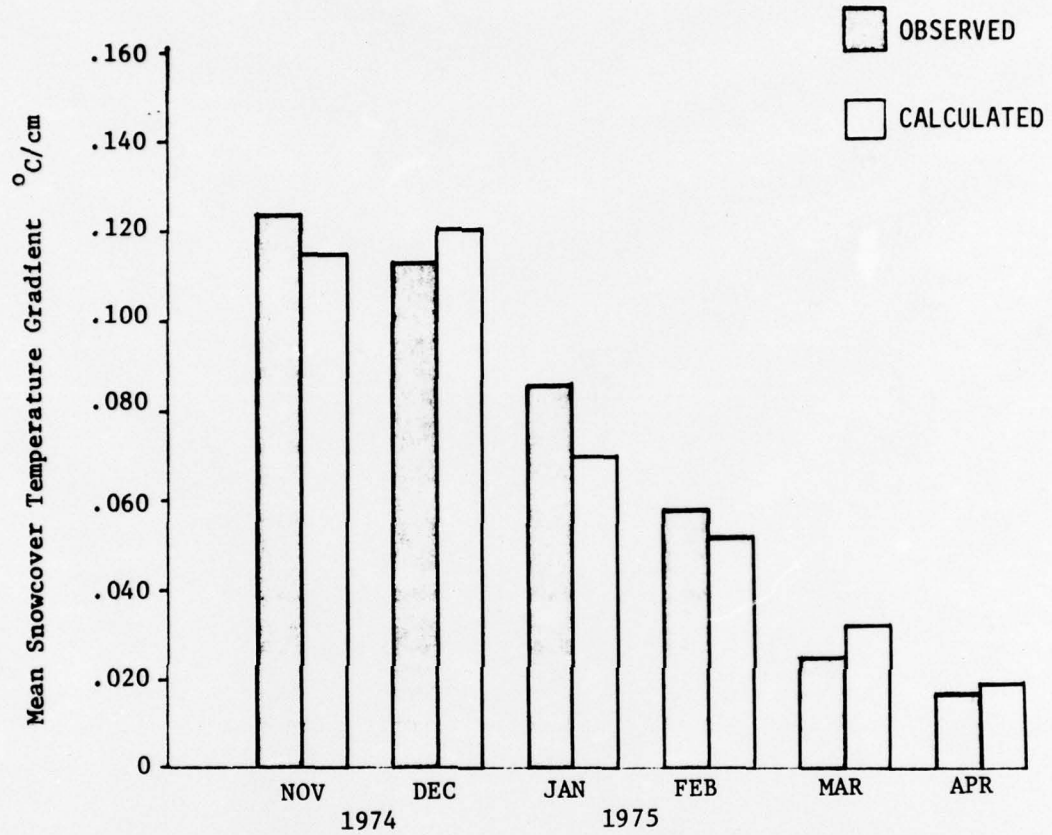
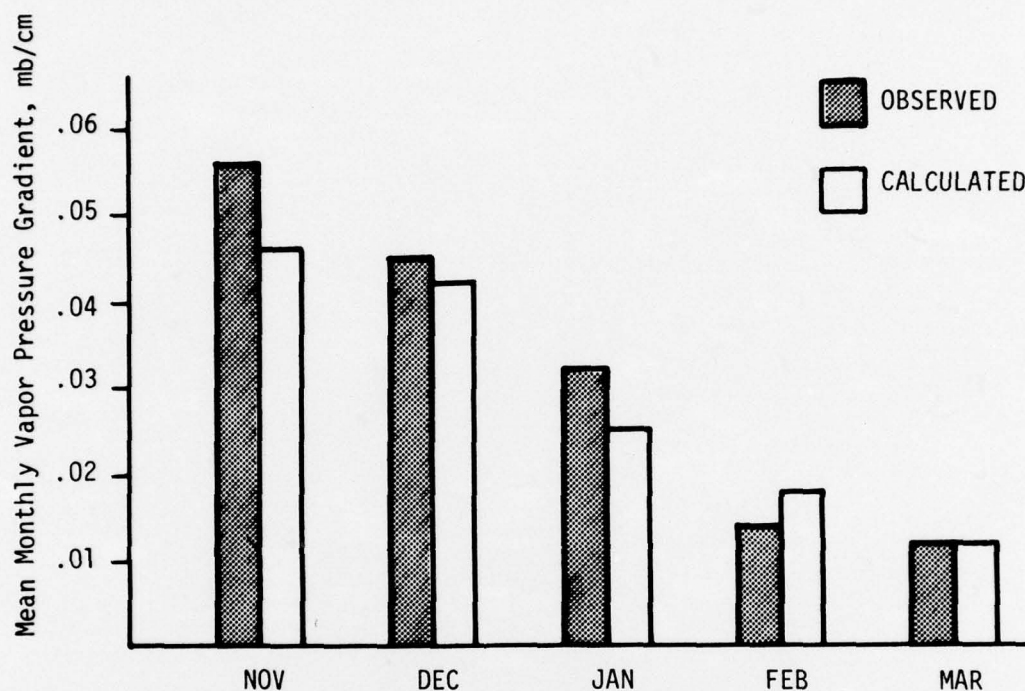


Figure 13 A comparison of observed and calculated mean monthly temperature gradients through the entire snow cover at Red Mountain Pass for the winter of 1974-75. The calculated values were obtained by dividing mean monthly air temperature by mean monthly snow depth according to the method of Akitaya (1974).



**Figure 14** A comparison of observed and calculated vapor pressure gradients (mean) through the entire snow cover at Red Mountain Pass for the winter of 1974-75. The method for calculating the gradients is explained in the text.

in both November and December, when pronounced TG metamorphism was observed, but in Figure 14 the critical vapor pressure gradient of 0.05 mb/cm is exceeded by the apparent monthly snow cover mean only marginally in November. In fact, a substantially larger vapor pressure gradient existed in the lower snow layers during these two months when TG metamorphism prevailed.

The effect of this non-linearity is exhibited in Table 3, where for the sake of uniformity a hypothetical snow cover of  $H = 100$  cm is considered exposed to different surface (air) temperatures while the snow-earth interface remains at  $0^{\circ}$  C. The temperature gradient is assumed to be uniformly at the mean value throughout the snow cover. The mean vapor pressure gradient is then calculated for the entire snow cover and for the bottom 10 cm of the snow cover. These two gradients diverge in value as the surface temperature falls. When the mean snow cover gradient reaches the critical value of 0.05 mb/cm, the gradient in the bottom 10 cm is already almost twice as great.

It thus appears that as a first approximation the conditions critical to TG metamorphism can be determined from mean air temperatures and snow depths as long as it is understood that comparison of the calculated temperature gradients with the nominal critical value of  $0.1^{\circ}$  C is strictly applicable only for those snow layers closest to the ground and to  $0^{\circ}$  C in mean temperature. Mean monthly vapor pressure gradients for an entire snow cover are less useful for identifying TG metamorphism conditions. Reasonable assumptions can be made about temperature gradient distribution within the snow and the local mean vapor pressure gradients deduced for individual snow layers.

Table 3

Comparison of saturation water vapor pressure gradients in a 100-cm snow cover for the two snow layers  $H = 0-10$  cm and  $H = 0-100$  cm (entire snow cover) for different values of snow surface temperature. Snow-earth interface temperature is assumed to be  $0^{\circ}\text{C}$ .

Snow surface temperature $^{\circ}\text{C}$	Mean temp. gradient $^{\circ}\text{C}/\text{cm}$	Vapor pressure gradient	
		0-10 cm mb/cm	0-100 cm
-2	0.02	0.010	0.009
-5	0.05	0.025	0.021
-8	0.08	0.039	0.030
-10	0.10	0.048	0.035
-15	0.15	0.071	0.045
-20	0.20	0.093	0.051
-25	0.25	0.115	0.055

### Bibliography

- Akitaya, E. 1974. Studies on Depth Hoar. Institute of Low Temperature Science, Hokkaido University, Sapporo, Japan, Series A, No. 26, 67 pp.
- Armstrong, R. L. and Ives, J. D., eds. 1976. Avalanche release and snow characteristics. University of Colorado INSTAAR Occasional Paper No. 19, 256 pp.
- Bradley, C. C., Brown, R. L., and Williams, T. 1976. Gradient metamorphism zonal weakening of the snowpack and avalanche initiation. Symposium on Applied Glaciology, September 1976, Cambridge, England. (in press)
- Bucher, E. and others. 1940. Lawinen, die Gefahr für den Skifahrer. Geotech. Kommission der Schweiz. Naturforschenden Gesellschaft.
- Eugster, H. P. 1952. Beitrag zu einer Gefügeanalyse des Schnees. Beitr. z. Geologie der Schweiz, Geotech. Serie, Hydrologie, 5. Lieferung.
- LaChapelle, E. R. 1974. In: Armstrong, R. L., LaChapelle, E. R., Bovis, M. J., Ives, J. D. Development of methodology for evaluation and prediction of avalanche hazard in the San Juan Mountain area of southwestern Colorado. University of Colorado INSTAAR Occasional Paper No. 13, 141 pp.
- Mellor, M. 1974. A review of basic snow mechanics. International Symposium on Snow Mechanics, Grindelwald, Switzerland, April 1974.
- Pahaut, E. 1975. Les cristaux de neige et leurs métamorphoses. Monographie No. 96 de la Meteorologie Nationale, Centre d'Étude de la Neige, St. Martin d'Hères.
- Sommerfeld, R. A. and E. R. LaChapelle. 1970. The Classification of Snow Metamorphism. J. of Glaciology, Vol. 9, No. 55, pp. 3-17.
- Trabant, D. and Benson, C. 1972. Field experiments on the development of depth hoar. Geological Society of America Memoir 135, Geophysical Institute of Alaska Contributions Series A388, College, Alaska, pp. 309-322.